A COMPARISON OF FLOW INTENSITIES IN ALLUVIAL RIVERS: CHARACTERISTICS AND IMPLICATIONS FOR MODELLING FLOW PROCESSES

NICHOLAS J. CLIFFORD

Department of Geography and Jackson Environment Institute, University College London, 26 Bedford Way, London WC1H 0AP, UK

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ABSTRACT

This paper compares flow intensity data obtained with different instruments from a variety of fluvial environments. It examines associations between the root-mean-square of longitudinal velocity fluctuations (flow intensity), local mean velocity, relative depth, and boundary resistance. Results indicate systematic differences in the behaviour of flow intensity which scale with respect to position in the boundary layer (deep sand-bedded rivers), boundary grain resistance (shallow river environments with coarse beds), and possibly form resistance (shallower sand-bedded rivers). Preliminary approaches to prediction and modelling of variations in flow intensity are suggested based upon linear regression relationships. Intensity values are also compared with theoretical and empirical limits to the use of Taylor's substitution, which allows time and frequency properties of a single-point velocity time series to be used to yield a flow length scale. In general, limits are exceeded in all environments for near-boundary flow measurements, but are met for $y/d > 0{\cdot}3$ in most cases in sand-bed rivers, and for $y/d > 0{\cdot}4$ in some gravel-bed environments. © 1998 John Wiley & Sons, Ltd.

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INTRODUCTION

Investigation of the character of small-scale fluid fluctuations in natural geophysical boundary layers has been a subject of considerable and growing interest during the last 25 years. From a theoretical standpoint, most attention has focused on comparisons with laboratory-derived 'burst–ejection' models of flow structure (for review see Clifford and French, 1993a, b; Best, 1993), while from a practical standpoint, the primary concern has been to provide greater insight into the initiation of particle entrainment (e.g. Drake *et al*., 1988; Williams *et al*., 1989b) and, to a lesser extent, to link bedform hierarchies with observations of coherent flow structures (e.g. Jackson, 1975, 1978). Most recently, attention has also turned to the need for more field measurements to inform closure assumptions in two- and three-dimensional numerical flow simulations. Despite increasing efforts, however, field studies are still at an early stage of development, and lack both the practical and conceptual sophistication of their laboratory counterparts. Apart from mean velocities, flow intensities are rarely measured in the field, and the distributional characteristics of fluctuating velocity components are frequently 'reconstructed' on the basis of laboratory observations used in conjunction with time-averaged phenomenological models such as the logarithmic velocity law (e.g. Bridge and Bennett, 1992; Kelsey *et al*., 1994).

A further problem concerns the identification and parameterization of small-scale flow structure, which can be associated with the roughness lengths and eddy viscosities used in hydraulic models (Clifford *et al*., 1992). Whereas laboratory studies routinely employ flow visualization and spatial correlation approaches made possible by dense instrument arrays, in field investigations most emphasis has been placed on interpretation of single-point or a few point measurements. Field instrument arrays are too expensive and/or cumbersome to deploy at small, tightly controlled separations and, in these circumstances, an alternative methodology based upon autocorrelation is used. Autocorrelation approaches are especially useful because they can be used with measurements of mean velocity to yield length-scale estimates of individual flow structures and to convert

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frequency spectra to wavenumber spectra for assessment of turbulent energy decay from which shear stress may be calculated. Nevertheless, the ability to use these methods successfully depends upon low levels of flow intensity, and an approximation to uniformity of the flow field.

In river environments, order-of-magnitude variations in depth:grain size ratio, multiple shear layers and intense, poorly specified combinations of vortex shedding and jet-and-wake processes are common. There is, therefore, a clear need for some very basic field data which can be used with theoretical and semi-empirical approaches to: (a) model variations in flow intensity with position in the boundary profile and/or with varying boundary characteristics; and (b) assess the limits of the applicability of autocorrelation methodology. The purpose of this paper is twofold: first, to provide a synthesis of comparative data obtained with differing instrumentation over a range of river scales, relative roughness, discharge and bedform type; and second, to comment upon the significance of these data for modelling flow properties and for informing future research.

DATA SOURCES AND METHODS

This paper brings together data from four studies. The two most extensive sets of observations are McQuivey's measurements (McQuivey, 1973) and those of Clifford (1994). McQuivey's data were obtained using hot-wire anemometers sampling at 100 Hz over 3 min in a 2·6 m wide flume with rectangular block roughness (2·7 mm high) and natural river sand ($d_{50}=0{\cdot}25$ mm, $\sigma=1{\cdot}44$ mm; McQuivey, 1973, pp. B17–B22 and table 5); and from measurements in natural channels (Mississippi, Missouri and Rio Grande conveyance channel) taken over dune bedforms with hot-film sensors sampled at 100 Hz over 4 min (McQuivey, 1973, pp. B12–B28 and tables 8, 10 and 11). Data reported in Clifford (1994) were obtained for contrasting width–depth and relative roughness conditions in a 200 m braid bar of Langden Brook, NW Lancashire, UK. Instrumentation consisted of discoidal (5·5 cm head) Valeport series 800 electromagnetic current meters (EMCMs) deployed at the channel centre of 10 cross-sections at 0·2, 0·4 and 0·8 relative depth over a range of flows corresponding to approximately 10–90 per cent duration. Centreline depths varied between 0·08 and 0·72 m, with channel widths of 2·5–9·5 m. Field sedimentological surveys (based on a 0·5 m grid and transect sampling following Wolman, 1954) revealed maximum particle sizes of 0·090 m throughout the measurement sections, but the d_{50} of the *b* axis varied from 0·055 to 0·070 m depending on the local subreach morphology. A total of 170 velocity time series were obtained, each sampled at 10 Hz over 3 min. Additional, less extensive data are presented from studies using a continuous output Ott dynamo current meter in the gravel-bed Squamish and North Alouette Rivers, British Columbia (Rood, 1980); with a Marsh-McBirney EMCM in the sand-bedded Squamish estuary (Babakaiff, 1993); and with a Colnbrook Mk11 EMCM (an earlier design of the Valeport series 800) in the gravel-bedded River Quarme in SW England (Cliford, 1990).

The methodology of the paper rests on the derivation and analysis of four flow properties. These relate to Reynolds' (1895) decomposition of a steady longitudinal flow velocity, U, at any given instant into the mean, $\overline{u}$ and fluctuating part, u, such that $U=u+\overline{u}$. The two most basic properties are the absolute and relative flow intensities. These are defined as the root-mean-square (RMS) of the longitudinal velocity, u', and this value relative to the mean longitudinal velocity, $u'/\overline{u}$. Attention is also given to two methods of deriving a macrolength scale characteristic of the flow field.

First, is the conventional application of the autocorrelation function:

$$r(t)=\frac{\overline{U(t)U(t+\tau)}}{u'u'} \tag{1}$$

which measures the correlation between $U(t)$ and $U(t+\tau)$, where τ is a fixed time step. Using Taylor's substitution (Taylor, 1938), in which distance, dx, is equal to the mean velocity multiplied by the time delay ($\mathrm{d}x=\overline{u}\mathrm{d}t$), an integral length scale of any turbulent structure is defined by:

$$L=\overline{u}\int_0^\infty r(t)\,\mathrm{d}t \tag{2}$$

(for examples of application, see West *et al*., 1986; Williams *et al*., 1989a; French and Clifford, 1992). L effectively represents the longest correlation distance between velocities at two points within the flow, and is termed the macroscale of turbulence. A similar argument is applied to the calculations of one-dimensional spectrum functions from frequency spectra of velocity fluctuations, used to define the microscale of turbulence of the flow (see McQuivey, 1973). Stapleton and Huntley (1995) present a recent extension to this where shear stresses are derived from the energy decay in the wavenumber spectrum. A more general review of the use of autocorrelation and spatial correlation approaches is given in Allen (1985).

Second, is a related technique based upon fitting pseudo-periodic stochastic autoregressive models to velocity time series. Generally, the autoregressive model describing the fluctuations in a velocity time series, u_t is defined by:

$$u_t = a_1 u_{t-1} + \ldots\ldots\ldots a_n u_{t-n} + e_t \tag{3}$$

wheref a_1–a_n are coefficients of the velocity at a given lag, and e_t is a random component. With respect to flow characterization, particular attention has been given to a subset of second-order (AR(2)) models. These involve only two coefficients and time lags, and are of special importance since, provided that $a_1^2 < -4a_2$, the AR(2) model describes a strictly pseudo-periodic process, the 'average' frequency, f_0, of which is defined as (Jenkins and Watts, 1968, p. 166):

$$\cos 2\pi f_0 = a_1 / 2\sqrt{-a_2} \tag{4}$$

The pseudo-periodic model provides a good description of series which characteristically yield broad-peaked spectra (Harvey, 1993). Because only certain series are expected to meet these criteria, the technique is particularly interesting, because it can be used both to suggest a significant degree of flow coherence, and as an indicator of the general dimensions of this coherence specified by the parameters of the model. Clifford *et al*. (1992) and Robert *et al*. (1993) provide illustrations of how the method can parameterize roughness lengths related to various scales of channel boundary resistance, and can also help to identify larger coherent flow structures where various poorly specified flow structure 'signals' combine in a 'noisy' flow record. Further discussion of the identification of appropriate AR models in the context of flow characterization is given in Clifford (1993).

When calculating length scales as described above, it is important to assess the degree to which the flow field satisfies the theoretical conditions upon which autocorrelation methods rely. Physically, Taylor's substitution (Equation 2) requires strict ergodicity: the sequence of events at a fixed point in the turbulent flow is assumed equivalent to the movement of an unchanging pattern of turbulence past that point (the so-called 'frozen flow assumption'; Townsend, 1980, pp. 64–65). Fluctuations at a fixed point may be imagined as caused by the whole turbulent flow field passing that point with a constant velocity. Velocity fluctuations at the point will, therefore, be almost identical to the instantaneous distribution of the velocity along the longitudinal axis through that point (Hinze, 1959, p. 51). For this to obtain, instantaneous fluctuations must be small compared to the local time-averaged velocity about which they are measured, and the eddy shape responsible for fluctuations should not be significantly altered during the time interval for movement past a measuring point (Bradshaw, 1971, p. 30). As Townsend (1980) explains, any observed decrease in maximum autocorrelation with time is assumed to arise mostly from the random movements of eddy centres. It is only if the random displacements are small compared with eddy diameters that the changes in the autocorrelation are produced by the *mean* velocity. Successful application of Equation 2 rests, therefore, on a close approximation to uniformity of the main flow and a low level of turbulence, which is frequently the case in laboratory situations. Before time–space substitution is used in the field, however, an assessment must be made of its applicability in the particular circumstances of observation. In this paper, two criteria are employed. First, as an overall indicator of the limits to the applicability of Equation 2, a relative flow intensity value of 0·1 is used. This follows standard engineering practice for *grid* turbulence (Reynolds, 1974, p. 91). The second criterion draws upon work

specifically relating to shear flows, and involves the comparison of maximum eddy dimensions in the flow to the magnitude of local velocity gradients.

Lin (1953) has shown that, theoretically, strict application of Equation 2 cannot generally be sustained in shear flows. Large eddies in the fluid motions are inevitably distorted by the shearing motion while they are carried past a fixed point, and eddy displacement is a function of *local* velocities caused by the larger eddies and instabilities created by eddy decay (Kaplan and Dinar, 1988). In these circumstances, a variety of limits can, however, be calculated which relate to the ability of specific components of the frequency spectrum of the turbulence (effectively, a lower frequency limit or maximum eddy dimension) to satisfy alternative criteria.

Generally, outside conditions of strict uniformity of flow, Equation 2 holds where:

$$U\frac{\delta U}{\delta x} \gg v\frac{\mathrm{d}U}{\mathrm{d}y} \tag{5}$$

and for components of the flow with wavenumbers k such that (Lin, 1953, p. 305):

$$kU \gg \frac{\mathrm{d}U}{\mathrm{d}y} \tag{6}$$

Using Equations 5 and 6 in conjunction with measured velocities and shear, it is possible, therefore, to estimate the maximum eddy size to which Equation 2 can be applied, and to assess the significance of this limitation with respect to the frequency or period (p) of the shear-contributing fluctuations which are present in any given environment. Appropriate comparisons depend upon the quality of measurements available and the variability of the flow in the vertical and horizontal. Using data from Townsend (1951) together with their own measurements obtained in laboratory shear layers, Klebanoff and Diehl (1951) suggest the following limits where shear varies as a function of the relative depth:

$$p < 0.7\frac{d}{U} \qquad \text{at } y/d = 0{\cdot}2 \tag{7}$$

$$p < 4.0\frac{d}{U} \qquad \text{at } y/d = 0{\cdot}4 \tag{8}$$

In Equations 7 and 8, original wavelength criteria have been converted to periods. Periods are specified because, in the absence of spatial correlation results from river environments, the only estimate of the size of coherent flow structures necessarily derives from examination of the time series properties of the flow from single-point measurements.

RESULTS AND DISCUSSION

Flow intensities obtained in contrasting rivers and with different instruments

Relative flow intensities for the river data sets are shown with relative depth in Figure 1. Several points emerge from this comparison. First, it is clear that all rivers display an increasing relative intensity as the boundary is approached. Apart from measurements in the Missouri, the 0·1 limit identified above is also frequently exceeded, although for the other sand-bedded rivers (open symbols), most values do satisfy the limit where $y/d > 0{\cdot}3$. In contrast, it is only in the upper parts ($y/d > 0{\cdot}4$) of gravel-bed river profiles (solid symbols) that any observations below the 0·1 limit are observed. Second, gravel-bedded rivers exhibit higher overall intensities than sand-bedded rivers. Gravel-bedded rivers also display an enhanced gradient of intensity increase towards the bed. Third, for the two most extensive data sets, the Mississippi River and Langden Brook, there is also evidence of increasing scatter and positive skewness in the intensity values in the near-bed zone. For the three sand-bedded rivers and for the Squamish and Quarme data sets, relative intensity is almost

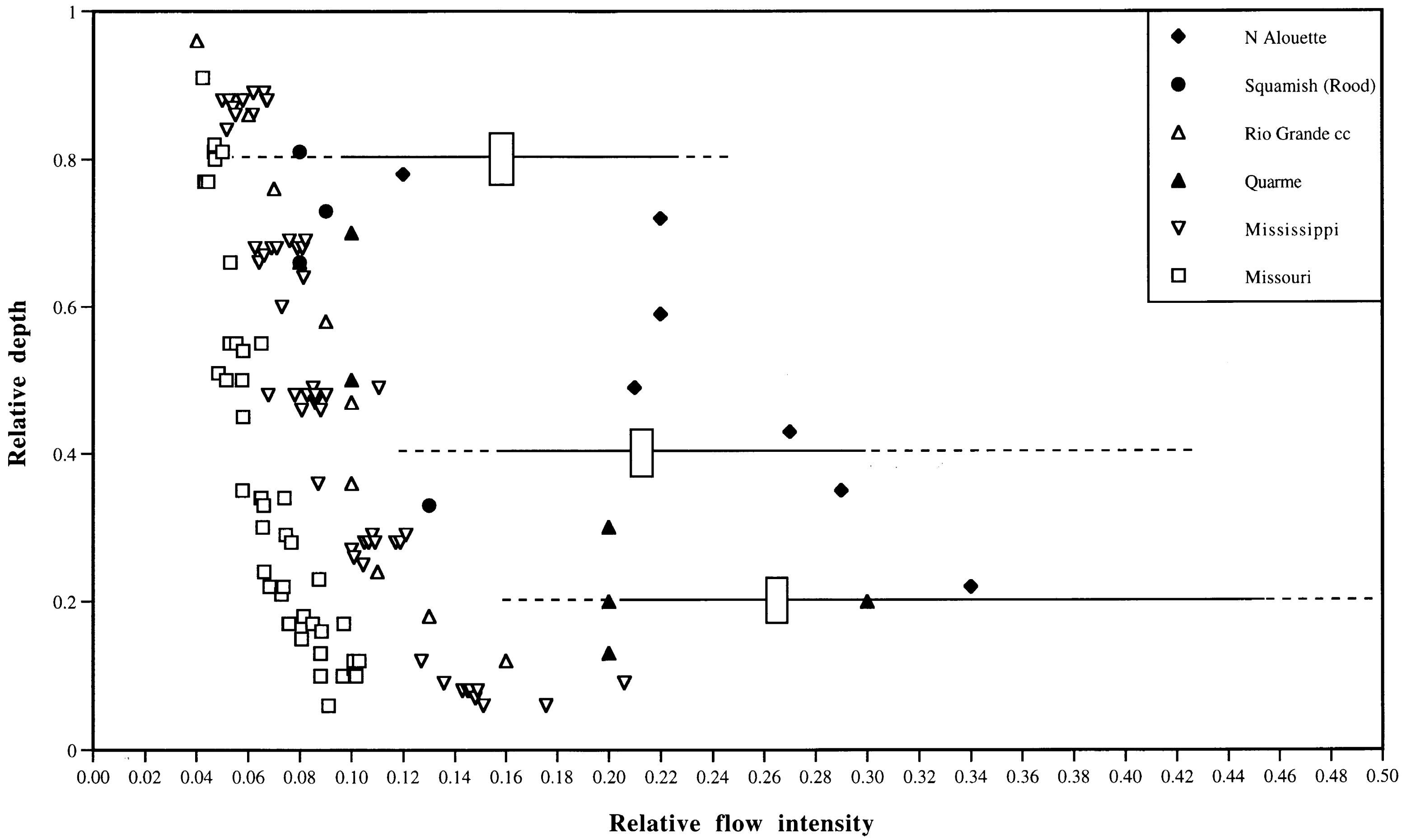

Figure 1. Comparison of relative flow intensities for contrasting rivers at varying relative depth. Where full boundary profile data exist, measurements are plotted with an individual symbol for each relative depth. For the case of the Langden Brook measurements, where all data were obtained at one of three fixed relative depths, the mean value for a given depth is plotted with a large rectangle, and the range of results indicated by solid (second and third quartiles) and dashed (minimum and maximum) lines

constant over the upper 60–70 per cent of the profile. After some consideration of the importance of flow sensor characteristics, the reasons for these complexities are examined below.

The role of instrument response. Because the gravel-bedded data were derived from instruments with much lower response times than the sand-bedded data, the data in Figure 1 should be interpreted with respect to the possible influences of instrument type and response rate on the range of values observed. In particular, impeller current meters are expected to pick up larger scale and slower flow excursions rather than higher frequency components of flow variation. As a result, systematic variations in the degree of comparison between intensity measures and flow scale relating to current meter versus electromagnetic or anemometer data might, therefore, be evident. McQuivey (1973, tables 12a–c) produces three sets of comparative data derived from impeller current meters and hot-film anemometers in the Mississippi, Rio Grande conveyance channel and the Atrisco feeder canal. Of the 11 series of measurements, all but one demonstrate a coefficient of variation which is generally a factor of two to three lower when calculated on the basis of impeller current meter data. Clearly, therefore, instrument response can introduce significant variations in results. No comparative data are available to assess the effects upon calculations of flow length scales. This is unfortunate, since intuitively, the relatively large inertia of water might be expected to reduce the relative importance of high frequency variations to measures of flow intensity, whereas measures of flow scale might be more sensitive (Church, pers. comm.).

Despite possible instrument effects, there are in fact several causes for optimism regarding the comparability of the data presented here. Even with the lower response times of the instruments (EMCMs) used in the gravel-bedded environments, results in Figure 1 very clearly demonstrate significant differences between sand- and gravel-bedded rivers. These may be interpreted with respect to contrasting boundary conditions (see below), and, if distorted in the manner suggested by McQuivey's results, the observed differences would be even more marked. It is interesting too, that, in the case of the coarse-bedded environments reported here, results from impeller (N Alouette) and electromagnetic current meter data (Quarme, Langden Brook) are comparable. The electromagnetic current meter data are also capable of yielding intensities close the the minimum recorded by the higher frequency measurements from the sand-bedded environments. These findings suggest, therefore, that while there may be a relationship between instrument response time and recorded RMS, this is probably constant above frequencies of approximately 5–10 Hz (i.e. the limiting response time of the gravel-bedded instrumentation reported here). It should also be noted that instrument response effects can be evaluated only when series length is sufficient to capture any significant elements of flow structure, with the further proviso that flow series have been appropriately detrended to take account of longer period phenomena which impart local flow unsteadiness (Bedford, 1991; Clifford and French, 1993b). All results presented here are based upon data series of very similar length, but a more extensive analysis of this, following Rood (1980), and of the effects of instrument type would obviously be desirable.

Relationships between flow intensity, mean velocity and local shear

Comparisons of absolute intensities against local mean velocities are given in Figure 2. Considering all data together, the major pattern is one of an overall increase in intensity with mean velocity, but the behaviour of gravel-bedded data and sand-bedded data is again distinguishable. Gravel-bed data sets (solid symbols) exhibit a tendency towards greater absolute intensities and there is also evidence of consistently greater scatter in these results compared to sand-bedded data sets. It is also the coarse-bedded data which, taken as a whole, show the steepest intensity:velocity relationship. More complex relationships must also be considered than a simple sand/gravel distinction, however. For example, data from the Missouri and Mississippi scatter conspicuously less than those from the Rio Grande conveyance channel or the Squamish estuary. This may reflect boundary layer disturbances arising from the relatively restricted depth compared to bedform height in the case of the conveyance channel (see below), and from the fact that the Squamish results in this case are tidally unsteady. Again, there is also a clear substructure to the overall pattern. Within data sets from the same river, there is a tendency for 'runs' of data which plot from upper left to lower right in the figures over most of the data range. Finally, in contrast to this, at the lowest mean velocities ($<c.\ 0{\cdot}25\,\mathrm{m\,s^{-1}}$), there is a very strong tendency towards a direct proportionality between mean and RMS values in the lower velocity regions of both the flume data set and data from Langden Brook, rather than the inverse depth-to-RMS relationship dominant in the other data

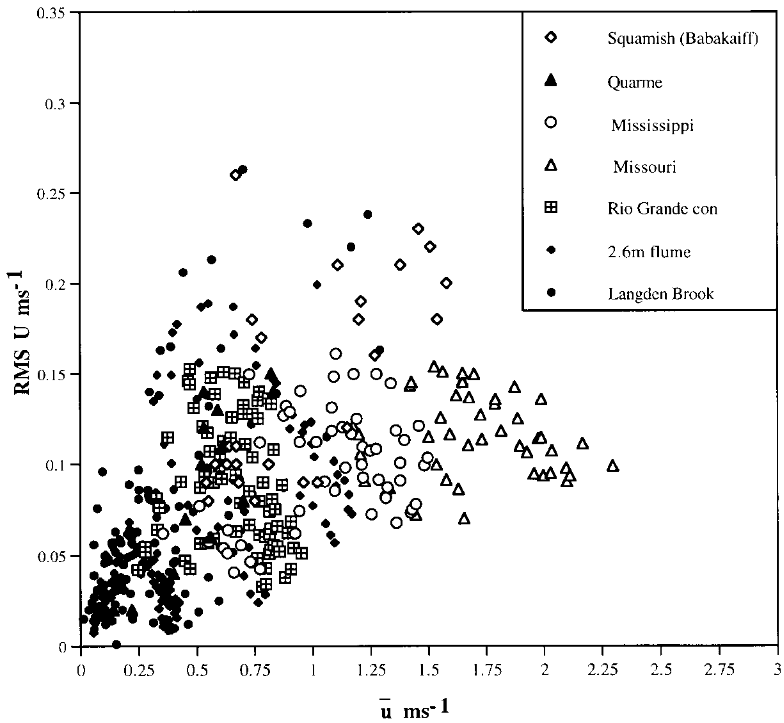

Figure 2. Comparison of RMS (absolute) flow intensities and local mean velocity for contrasting rivers

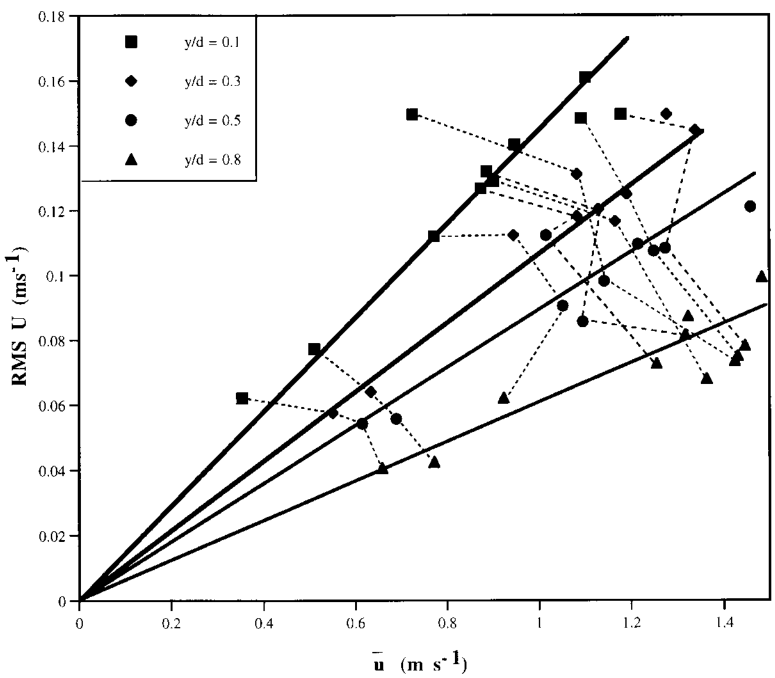

Figure 3. Possible linear relationship between absolute (RMS) flow intensity and local mean velocity for contrasting relative depth on the Mississippi River. Data from McQuivey (1973). Dashed lines show the velocity profiles from which relative height data were obtained

sets. Figures 3 and 4 are an attempt to rationalize the behaviour shown in Figures 1 and 2 with respect to the size and nature of the boundary roughness in relation to the depth of flow and structure of the boundary layer.

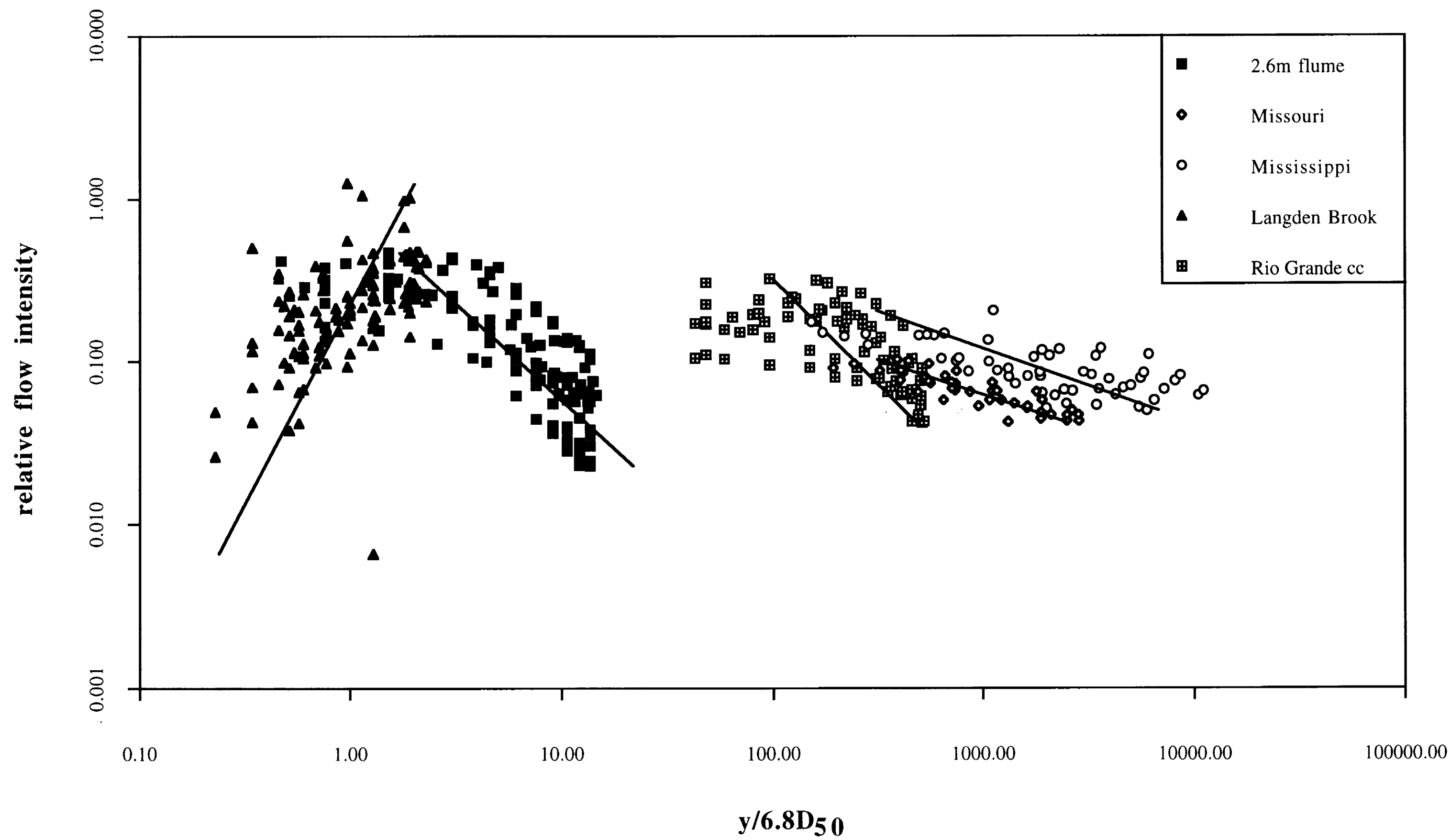

Figure 4. Comparison of relative flow intensities with relative roughness scale (Naden, 1981) for contrasting rivers

Table I. Linear regression coefficients describing relationships between RMS flow intensities and local mean velocity at varying relative depths in contrasting rivers and river reaches. Regressions were fitted without a constant term to reflect zero intensity at zero mean velocity. Figures in parentheses indicate standard deviation about the regression line

	y/*d* value			
River	0·1	0·3	0·5	0·8
Langden Brook	0·245 (0·054)	0·225 (0·048)		0·147 (0·022)
Mississippi	0·146 (0·016)	0·110 (0·008)	0·087 (0·008)	0·058 (0·009)
Missouri	0·095 (0·010)	0·067 (0·014)	0·055 (0·010)	0·045 (0·005)

In Figure 3, absolute intensities for the Mississippi have been plotted versus $\overline{u}$ at fixed relative depths (y/d) in order to control for the joint effects of increasing local mean velocity and boundary profile shear. All data exhibit two aspects of a dependence upon mean velocity. First, intensity increases for individual 'profiles' of data as overall mean velocity increases (dashed lines linking points for $y/d = 0{\cdot}1$–$0{\cdot}8$ running from top left to bottom right across the diagram). The range of intensity values in any given profile also increases with mean velocity. Second, there is an increase in RMS levels for individual measurements at fixed relative depth as mean velocity increases. There is sufficient regularity in the behaviour of data points at the same relative depth to attempt to model this variation using simple linear regression. The resulting models suggest that the greatest increases in intensity with local mean velocity occur closest to the bed. Regression coefficients are listed in Table I for straight-line fits to changes in absolute intensity with local mean velocity at varying relative depth such as those in Figure 3, but for various rivers. The two large sand-bedded rivers are closely similar in behaviour, while the results for Langden Brook show more sensitivity to local mean velocity, with regression coefficients which are two to three times larger for the same relative depth. Data from the sand-bedded rivers also exhibit much less scatter about the regression line than data from Langden Brook. The most obvious factor likely to explain observed intensity is the degree of shear in the boundary profiles. Both relative and absolute intensities are highest closer to the bed, and over rougher boundaries where shear is most intense. Gravel-bedded rivers and near-bed regions of sand-bedded rivers thus have the highest absolute and relative intensities because zones of higher shear are associated with zones of lower velocity *provided that* a well-developed boundary layer exists. Any relationship between intensity and mean velocity is, however, likely to be stage-dependent, too, especially in conditions of high relative roughness where very local factors such as position of measurements in or out of the immediate jet-and-wake regions around individual roughness elements will be the dominant control (Wang *et al*., 1993).

In an attempt to clarify the relative influence of boundary resistance on flow intensity, Naden (1987) presents a best-fit regression relating relative flow intensity to an (inverse) index of relative roughness for McQuivey's flume data with 4·4 mm bed material. The regression equation is:

$$\frac{\sigma_{\mathrm{u}}}{u_{\mathrm{y}}} = 0{\cdot}16\left(\frac{y}{6{\cdot}8D_{50}}\right)^{-0{\cdot}65} \tag{9}$$

Choice of an appropriate roughness parameter is somewhat arbitrary, but the $6{\cdot}8D_{50}$ (equivalent to $3{\cdot}5D_{84}$) has been found to improve mean velocity prediction based upon use of the logarithmic flow law in coarse-grained environments (e.g. Bathurst, 1982) and can further be physically justified with respect to the influence of grain microtopography and its effect on flow structuring (Clifford *et al*., 1992). Naden's result in Equation 9 may be compared with regressions of the same form detailed in Table II and shown on Figure 4. In this figure variations in relative flow intensity in the two deeper, sand-bedded channels are very clearly distinguished from the shallow, coarse-bedded channels, with the Rio Grande conveyance channel data lying somewhat intermediate between these. In the deep channels, relative intensity declines through a relatively restricted range with reductions in relative roughness of *c*. 200–10 000. In the shallow channels, relative intensity varies more, with

Table II. Regression coefficients, exponents and correlation coefficients describing relationships between relative flow intensities and relative roughness for contrasting rivers and flume data. Analysis follows Naden (1981)

Data set	Coefficient	Exponent	*r*
McQuivey 4·4 mm	0·16	–0·65	0·96
McQuivey 8 ft	0·91	–0·84	0·76
Missouri	1·58	–2·33	0·89
Mississippi	10·06	–2·17	0·71
Rio Grande c.c.	41·61	–0·77	0·56
Langden Brook	1·84	0·41	0·53

relative roughness varying from *c*. 0·2 to 10·0. Up to relative roughness values of approximately 1, intensity actually increases as relative roughness declines, but thereafter the general trend of declining intensity with declining relative roughness is established. Langden data show only the increasing trend, while the flume data span both 'regions' of behaviour, lying around the relative roughness value of 1·00. The conveyance channel data also resemble the flume data, although with relative roughnesses two orders of magnitude less. Thus, at relative roughness value of approximately 200, relative flow intensity begins to decline systematically as relative roughness declines, with no systematic behaviour below this value. In an analysis restricted to McQuivey's flume data sets, Naden (1981) observed that, although log–log relationships were consistently applicable, no single transformation could be found to eliminate scatter, differing slopes and regression constants in relationships relating flow intensity to relative depth. The results in Figure 4 can be viewed in exactly the same way. Taken together, results from Figures 3 and 4 suggest the following interpretations. Generally, where boundary layers are deeper, variation of absolute intensity against mean velocity is less marked and relative intensities are much less variable as shear is relatively constant (i.e. semi-logarithmic velocity distribution approximates). For depth-limited flows, such as Langden Brook and the flume data, however, intensities show a more simple increase with mean velocity since the local mean velocity is itself a closer representation of the scale of velocity fluctuations at any height above the bed. Local intensities are probably dominated by wake vortex activity around bed roughness elements throughout the flow depth, and the RMS of wake fluctuations varies as a linear function of velocity for a certain range of grain Reynolds number (Hille *et al*., 1983). This condition persists while depth and velocities are relatively low. As depth and velocity increase, then the boundary layer profile becomes better established and, once more, the pattern is one in which absolute intensities *decrease* with increasing velocity within any *one* data set, i.e. are greater closer to the bed (reflected in the dashed lines of Figure 3). Observed intensity behaviour in the Rio Grande conveyance channel suggests that this kind of explanation can be extended to include the effects of mobile dune bedforms where, like isolated obstacles such as grains and grain clusters, they extend to significant portions of the flow depth. Although no measurements of dune heights are available, McQuivey notes that the conveyance channel data were obtained in conditions where dune bedforms were present, and depths were almost constant at 1 m. In contrast, dunes were also noted in the Missouri and Mississippi, but here, depths varied from *c*. 3 m (Missouri) to 7–15 m (Mississippi). Further analysis of stage- and bar-form dependencies of high frequency velocity characteristics are given in Clifford (1996).

Implications for modelling flow processes

In any shear flow, eddies of considerable size are likely to be changed by the shearing motion during their passage past any given point (Lin, 1953). It is evident from Figures 2–4 that the magnitude of this change will vary with the nature of the boundary and position within the profile, both of which relate to the magnitude of shear experienced. Simple regression equations are capable of describing much of the observed variation in absolute and relative flow intensities observed over a wide range of river environments, but the steepness and, in some cases, the direction of the regression line are specific to the particular river in question. A knowledge of the range of relative roughness appears to be vital in interpreting results.

Returning to the question regarding acceptable limits for the Taylor substitution in Equation 2, Figure 5 illustrates periods calculated from time series obtained in Langden Brook. These are compared with the maximum permissible periods (i.e. equivalent to a maximum eddy size) which meet the restrictions of

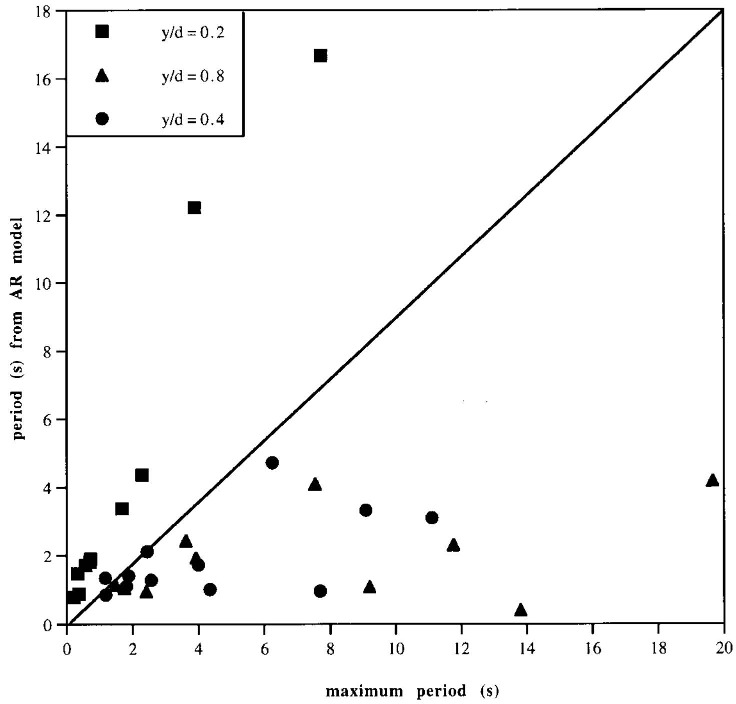

Figure 5. Comparison of periods describing possible flow structures in Langden Brook and maximum permissible periods for application of Taylor's time–space substitution

Equations 7 and 8. Limiting periods for relative depths (*y*/*d*) of 0·6 and 0·8 have been calculated assuming a linear extrapolation from values given by Equations 7 and 8. A line of equivalence is drawn, and points falling above this are 'unacceptable' for the Taylor substitution in Equation 2, while points below the line are 'acceptable'. In depth-limited gravel-bedded environments such as this, the results reveal again that restrictions are severe, and almost completely rule out the substitution close to the boundary. Beyond a mid-depth level, there is more cause for optimism. At 40 per cent depth, almost all series are acceptable, which is less restrictive than implied by Figure 1, when the standard engineering limit for a relative intensity of 0·1 is employed. Nevertheless, when interpreting these findings, it should be remembered that the limiting values are based upon reasonably well-specified shear flows derived under laboratory conditions. Less restrictive limits have been developed for isotropic aeronautical observations by Corrsin and Uberoi (1953), but the Langden measurements lack the detail required for comparison with their findings, and to derive the necessary velocity gradients given in Equations 5 and 6. Given the likelihood of embedded boundaries and jet-and-wake processes common to very rough gravel environments (Herbich and Schulits, 1964), more appropriate criteria are likely to be more, rather than less, restrictive and also more difficult to generalize than those employed here. Clearly, there is a requirement for further research in this respect. In the meantime, general application of time–space substitution should be treated very cautiously.

CONCLUSIONS

Empirical measurements of flow structure and velocity variation in a variety of fluvial environments suggest that both sand- and gravel-bedded rivers exhibit systematic relationships involving absolute and relative flow intensities, local and boundary profile mean velocities, and boundary profile shear. At constant relative depth within a boundary profile, intensities increase as local mean velocity increases, and this increase is greatest closer to the boundary. In contrast, for a fixed profile mean velocity, absolute intensities fall as local velocity increases and shear decreases away from the bed. Although the *nature* of these relationships is similar, sand-

and gravel-bedded environments can be distinguished by the *strength* of the relationships. Results from gravel-bedded rivers demonstrate more scatter than sand-bedded ones. For depth-limited flows where $y/6{\cdot}8D_{50} < c.\ 1$, there is some suggestion that a simpler relationship between flow intensity and mean velocity exists. In these circumstances, boundary layer profiles are poorly developed and flow behaviour is dominated by vortex shedding and localized jetting around roughness elements. Relative flow intensity thus increases with local mean velocity.

Use of Taylor's time–space substitution has been a popular method for connecting the time spectrum of flow velocities with the space correlation of flow structures in a wide variety of detailed flow processes studies in natural environments. For this to apply, however, the spatial pattern of turbulent motion is assumed to be carried past a fixed point by the mean flow without any essential change. For laboratory studies of grid turbulence, where flow is uniform and the level of turbulence is low, the assumption is satisfied. For shear flows, the assumption can never strictly be correct. If conventional engineering values specifying the limiting relative intensity for appropriate time–space substitution are used, data presented here demonstrate that, at low relative depths ($y/d < c.\ 0{\cdot}2$), use of Taylor's substitution is inappropriate in most river environments. Beyond relative depths of *c*. 40 per cent its use is likely to have greater justification, especially in sand-bedded rivers and in gravel-bedded rivers of lower relative roughness. Limiting values derived from laboratory shear layer experiments give greater cause for optimism, but these limits are probably generous, and there is a clear need for further field measurements to establish more appropriate values for use in natural geophysical boundary layers. For the present, general application of time–space substitution should, therefore, be treated with extreme caution.

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